Larsen Ice Shelf Has Progressively Thinned

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The retreat and collapse of Antarctic Peninsula ice shelves in tandem with a regional atmospheric warming has fueled speculation as to how these events may be related. Satellite radar altimeter measurements show that between 1992 and 2001 the Larsen Ice Shelf lowered by up to 0.27 ± 0.11 meters per year. The lowering is explained by increased summer melt-water and the loss of basal ice through melting. Enhanced ocean-driven melting may provide a simple link between regional climate warming and the successive disintegration of sections of the Larsen Ice Shelf.

On average, Antarctic Peninsula (AP) ice shelves have retreated by ~300 km² each year since 1980 (1). This gradual retreat has been punctuated by two catastrophic collapses, in January 1995 (2) and February 2002, when the remaining northern sections of the Larsen Ice Shelf (LIS) (Fig. 1) fragmented into icebergs. In contrast to the prolonged retreats, these 2000- and 3250-km² ice-shelf sections—Larsen-A and Larsen-B—disintegrated over days or weeks. Although the initial retreats of their ice fronts may have resulted from iceberg calving beyond stable geometrical positions (3), it is not clear that this explains the wholesale disintegration of large ice-shelf sections. Speculation as to the mechanism that caused the final collapses has concentrated on the destabilizing effects of increased surface melt-water (4–6), which may have enhanced the process of crevasse fracture (7). Although that mechanism provides a link between the regional climate warming and the breakup of ice shelves at the AP, direct observations are insufficient to determine the importance of ice-shelf stability criteria or the impact of increased surface melt. Here, we show that the LIS may have become susceptible to crevasse fracture through a sustained ice thinning.

We used 9 years of European Remote Sensing (ERS-1 and -2) satellite radar altimeter measurements to determine the surface elevation change of the LIS since 1992. Ice-shelf surface elevation was calculated relative to the WGS 84 Earth ellipsoid at each individual crossing point of the satellite ground track during 35-day orbit repeat cycles (8). From these data, we formed 45 time series of elevation change (e.g., Fig. 2) across the LIS at the finest resolution afforded by the ERS altimeters (9). The ERS ground tracks provided a mean crossing point separation of 14 km, and the precise location of the ~10-km footprint drifted by less than 1.2 km through time as a result of orbit maneuvers. We detected no radar penetration of the LIS surface at any time (10), and we removed the ocean tide from each elevation time series using predictions of ice-shelf tidal displacement (11). We calculated the trend of elevation change at each crossing point from the tide-adjusted time series and interpolated these data using a quintic triangulation scheme (Fig. 1).

Between 1992 and 2001 the mean rates of elevation change of the Larsen-B and -C ice shelves were −0.17 ± 0.11 and −0.08 ± 0.04 m yr⁻¹, respectively. In general, the northernmost sections of the Larsen-C experienced the greatest decrease in surface elevation, with a peak rate of lowering of 0.27 ± 0.11 m yr⁻¹ some 80 km west of the Larsen meteorological station (Fig. 1). Toward the southern tip of Larsen-C, one time series showed that the ice shelf thickened at the terminus of the Lubbeee Glacier (69.25°S, 63.62°W), where ice is discharged from Palmer Land. Recent radar, seismic, and Global Positioning System elevation surveys (12) show that sections of the Larsen-B lowered relative to the geoid by 0.18 m yr⁻¹ between 1991 and 1999, in agreement with our satellite-derived rate. The rate of elevation change of a floating ice shelf relative to the earth ellipsoid (Δh/Δt) is due to fluctuations in sea-level height (Δs), ocean (pₖ) and ice-shelf (pₛ) densities, net surface (Mₛ) and basal (Mₖ) mass accumulation, and ice-flux di-
vergences (\(M\dot{V},v\)). For an ice column in hydrostatic equilibrium with no vertical ice shear, \(\dot{h}/\dot{t}\) can be written as

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\frac{\partial h}{\partial t} = \frac{\partial \Delta s}{\partial t} - M \frac{\partial (\frac{1}{\rho_v})}{\partial t} + \int_0^h \frac{\partial}{\partial t} \left( \frac{1}{\rho_i} \right) dm, \quad (1)
\]

where \(\rho_v\) is the density of ice (917 kg m\(^{-3}\)) and \(M\) is the ice-shelf mass per unit area. We investigated trends and variability in each component of Eq. 1 to determine the origin of the LIS surface lowering.

External contributions to \(\dot{h}/\dot{t}\) include sea level and density. Although eustatic sea level has risen by only 2 mm yr\(^{-1}\) in the 20th century (13), atmospheric pressure fluctuations occur (14) that could lead to a 9-year uncertainty in \(\Delta s\), of up to 11 mm yr\(^{-1}\). Tide model inaccuracies introduce a further 26 mm yr\(^{-1}\) uncertainty. Oceanographic records show that seasonal and interannual variations in salinity (15) modify \(\rho_v\), resulting in up to 10 mm yr\(^{-1}\) uncertainty in elevation change. In addition, nearby deep and surface layers of the Weddell Sea have warmed by 0.01°C yr\(^{-1}\) for several decades (16, 17), and a similar warming (or cooling) of the water beneath the LIS could raise (or lower) the ice shelf at a rate of 15 mm yr\(^{-1}\). We estimate the combined uncertainty resulting from these contributions to be 34 mm yr\(^{-1}\), a value that is small when compared with the observed trend. The LIS lowering must reflect a change in the ice shelf itself.

We investigated the possibility that the LIS \(\dot{h}/\dot{t}\) may have resulted from a mass-conservative densification of the ice shelf (the third term in Eq. 1). The gradually warming (18) and lengthening (19) summer climate may have accelerated firm densification during the interval of the measurement. To bound this acceleration, we estimated an initial LIS firm density (20) and supposed that the entire upper 8 m of firm layer (beneath which interannual temperature changes do not penetrate) (12) were densified to ice. This calculation leads to rates of elevation change of \(-0.12\) and \(-0.28\) m yr\(^{-1}\) at the Larsen-B and -C, respectively, the difference reflecting the lower initial density of the colder Larsen-C. These changes are comparable to the observed lowering (Fig. 1), and densification cannot be excluded as the sole explanation of the lowering. However, the energy available for thermally driven firm densification is limited by air temperature (1) and melt-season duration (19), which decrease southward. We used observed ablation measurements and a positive degree-day model (21) to estimate the change in summer melt-water production for several decades that observed more precisely by the satellite. Allocating this rate to densification, for which the thinning rate equals the elevation rate relative to sea level, appears incompatible with the trend in melt-water production (22). Our conclusion is that, for Larsen-C at least, some other cause of thinning is present as well.

The remaining terms in Eq. 1 are mass losses. A 10% century\(^{-1}\) increase in snow accumulation (\(M\)) similar to that recorded nearby at the ~2000-m altitude Dyer Plateau (25) would increase surface elevation by some 2 mm yr\(^{-1}\). Fluctuations occur at shorter time scales (8), but snow-pit measurements at Larsen-B show that accumulation has remained close to

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**Fig. 1.** Rate of surface elevation change of the Larsen Ice Shelf determined from ERS radar altimeter measurements recorded between 1992 and 2001 (color scale and 0.1 m yr\(^{-1}\) white contours). The elevation data are superimposed on a mosaic of Advanced Very-High-Resolution Radiometer (AVHRR) satellite imagery (gray scale) (38), and data points used in the interpolation are shown as black dots. Also shown are the locations of meteorological stations (blue dots) and stakes used to measure changes in snow height (red dots) and surface mass balance (5-km grid, centered at green dot). The 1990 boundaries of the Larsen-A, -B, and -C ice-shelf sections are highlighted with blue, green, and red borders, respectively. Larsen-A collapsed before the ERS measurements; Larsen-B has since disintegrated; Larsen-C remains intact.

**Fig. 2.** Change in surface elevation between 1992 and 2001 recorded by the ERS radar altimeters ~80 km west of the Larsen meteorological station (see Fig. 1) at the Larsen-C ice shelf before (A) and after (B) removal of the periodic signal of ocean tide (11). On average, the tide correction reduced the variability in the trends from 0.4 to 0.2 m yr\(^{-1}\). The location of this time series is highlighted with a white border in Fig. 1.
Although basal melting rates under the LIS are uncertain and their fluctuation even more so, the conclusion that basal melting has been and is thinning the LIS, at an average rate our measurement and estimate of the effect of densification (Fig. 3) put at 0.19 m yr$^{-1}$ (29), but substantially greater melt occurs in regions where warm waters are transported beneath floating ice (30). Basal melt rates of 2 to 3 m yr$^{-1}$ are observed up to 200 km inshore of the FRIS ice front, where tidal mixing occurs (29). On average, nearby Weddell Sea Deep Waters (WDW) have warmed by 0.32$^\circ$C since 1972 (16). In 2002, oceanographic measurements showed large quantities of modified WDW present in front of the northern Larsen-C (66.5$^\circ$S) at depths well below the ice-shelf draft (300 m), with a potential temperature of $-1.45$ $^\circ$C, that is, 0.65$^\circ$C higher than the pressure melting point of ice (31). Elsewhere such differences generate up to 6.5 m yr$^{-1}$ of basal ice melt (30) when water is delivered to the ice-shelf base.

The calving front of the Larsen-C has changed little in decades, and its flow geometry is considered to be stable (3). However, at the same time enhanced ocean melting has progressively thinned the shelf at its base. Thinning inevitably increases the ice-shelf exposure to crevasses fracture (7), and it is difficult to conclude that the thinning did not contribute toward either the creation of unstable conditions or the final disintegrations of the Larsen-A and -B sections. If our estimate of the basal erosion rate is correct, the Larsen-C will approach the thickness of the Larsen-B at the time of its collapse in some 100 years, more rapidly if the rate is increased by a warming ocean. It is possible that the LIS thinning provides a link between the regional climate warming and the disintegration of ice shelves at the Antarctic Peninsula.
The growth of continental-scale ice sheets extending to the tropics during the second half of the Neoproterozoic (1000 to 540 million years ago) is now widely accepted in the geological community and has been of particular interest because of its close stratigraphic association with the first appearance of metazoans and the possibility that ice ages served as an environmental filter for animal evolution. The severity of these ice ages, which may record the coldest times in Earth history (3), implies that the Precambrian climate system must have operated very differently from today. This is supported by the ubiquitous occurrence of thin post-glacial “cap” carbonate units (4–7), apparent perturbations of the carbon cycle that did not recur in the Phanerozoic. To account for these observations, we focus on a first-order difference between the Precambrian and modern Earth systems and its implications for atmospheric CO$_2$; the absence of a well-developed deep-sea carbonate sink before the proliferation of calcareous plankton.

On the time scale of glaciations (~10$^4$ to 10$^6$ years), the balance between weathering of terrigenous rocks and the burial flux of calcium carbonate (CaCO$_3$) in marine sediments exerts a key control on ocean carbonate chemistry (8), with this burial today divided roughly equally between deep-water (pelagic) and shallow-water (neritic) zones (9). The latter sink is of particular relevance in the context of ice ages, because the total neritic area available for CaCO$_3$ burial is highly sensitive to sea level, a consequence of the nonuniform distribution of the Earth’s surface area with elevation (Fig. 1). The climatic relevance arises because any increase in the carbonate ion concentration ([CO$_3$$^2-$/HCO$_3^-$]) at the ocean surface will induce lower atmospheric CO$_2$ (because the aqueous carbonate equilibrium, CO$_2$ + CO$_3$$^2-$ + H$_2$O ↔ 2HCO$_3^-$, is shifted to the right). This is the basis for the coral reef hypothesis for Quaternary glacial-interglacial CO$_2$ control (10–13), in which lowered sea level reduces available neritic area and CaCO$_3$ accumulation rates, driving higher [CO$_3$$^2-$/HCO$_3^-$] and lower CO$_2$.

We have identified a fundamental difference between ancient and modern carbon cycles in the relative importance of the neritic carbonate sink that would make the impact of a coral reef–like effect much greater in the Precambrian. In the modern system, higher [CO$_3$$^2-$] enhances the preservation of carbonate in deep-sea sediments, hence, a reduction in neritic carbonate deposition due to a fall in sea level can be compensated for by a greater burial flux in deep-sea sediments of CaCO$_3$ that originates from planktic calcifiers (9) (Fig. 1). This provides a strong negative (stabilizing) feedback on the modern carbon cycle, restricting oceanic [CO$_3$$^2-$] variation and thus limiting the atmospheric response to sea level change.

The Neoproterozoic carbon cycle, by contrast, did not possess this stabilizing feedback, because before the advent of pelagic calcifiers in the Cambrian and the subsequent proliferation of coccolithophores and foraminifera during the Mesozoic (14), carbonate deposition would have been largely limited to neritic zones. The importance of the calcareous plankton that dominates carbonate deposition in the modern open ocean (9) is illustrated by the comparative rarity of deep-sea pelagic carbonate material in ophiolite suites older than ~300 million years (14). As neritic carbonate deposition was the dominant mechanism of CO$_3$$^2-$ removal in the Precambrian ocean, it follows that atmospheric CO$_2$ would have been much more sensitive to sea level change. We explore the implications for the Neoproterozoic carbon cycle of sea level variation with the aid of a numerical model (15). This model calculates the evolution in atmospheric CO$_2$ that arises from a reduction in the area available for neritic carbonate deposition.

Although these observational and evolutionary arguments suggest a highly limited role for the deep-sea carbonate buffer in the Precambrian carbon cycle, Precambrian ocean chemistry would instead have been stabilized by the dependence of shallow-water carbonate deposition rates on [CO$_3$$^2-$] (8). As oceanic [CO$_3$$^2-$] (and saturation state, $\Omega$) rises after a fall in sea level, the smaller area available for carbonate deposition is eventually compensated for by a higher precipitation rate per unit area. An analogous compensating increase in the neritic CaCO$_3$ precipitation rate may have occurred at the Cretaceous/Tertiary boundary after the extinction-driven reduction of pelagic carbonate productivity (8). The precipitation rate of carbonate minerals is expressed in the model as a proportionality with $\Omega$ (~1$^{-n}$) (16), where $\Omega$ is defined as ([Ca$^{2+}$]$^n$ × [CO$_3$$^2-$])$/K_{sp}$ (where $K_{sp}$ is a solubility constant). The parameter $n$ is a measure of how strongly CaCO$_3$ precipitation rate responds to a change in ambient [CO$_3$$^2-$], and thus of how effectively ocean chemistry and atmospheric CO$_2$ are buffered. Possible values range from ~1.0 for modern biological systems such as corals (17) to 1.9 < $n$ < 2.8 for precipitation that occurs under entirely abiotic conditions (16). We therefore initially set $n = 1.7$ (8, 11). Because CaCO$_3$ precipitation dur-